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Key Points:
- The *wet gets wetter, dry gets drier*, idea does not apply regionally, because stationary-eddy circulations shift as the climate warms
- The zonal variance of the hydrological cycle increases with global warming
- The increase in zonal variance of the hydrological cycle is limited by a weakening of stationary-eddy circulations

Supporting Information:
- Supporting Information S1

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Thermodynamic and dynamic controls on changes in the zonally anomalous hydrological cycle

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Abstract The *wet gets wetter, dry gets drier* paradigm explains the expected moistening of the extratropics and drying of the subtropics as the atmospheric moisture content increases with global warming. Here we show, using precipitation minus evaporation ($P - E$) data from climate models, that it cannot be extended to apply regionally to deviations from the zonal mean. Wet and dry zones shift substantially in response to shifts in the stationary-eddy circulations that cause them. Additionally, atmospheric circulation changes lead to a smaller increase in the zonal variance of $P - E$ than would be expected from atmospheric moistening alone. The $P - E$ variance change can be split into dynamic and thermodynamic components through an analysis of the atmospheric moisture budget. This reveals that a weakening of stationary-eddy circulations and changes in the zonal variation of transient-eddy moisture fluxes moderate the strengthening of the zonally anomalous hydrological cycle with global warming.

1. Introduction

The availability of water will be a crucial issue for society during the next century. It depends on the spatiotemporal variability of net precipitation (precipitation minus evaporation, $P - E$). In the absence of changes in atmospheric circulations, the change of $P - E$ with warming is simply related to the change in moisture content of the atmosphere [*Mitchell et al.*, 1987; *Chou and Neelin*, 2004; *Held and Soden*, 2006]. The moisture content of the atmosphere increases substantially with warming, approximately following the Clausius-Clapeyron relation, because of energetic constraints on relative humidity near the surface [*Held and Soden*, 2000, 2006; *Schneider et al.*, 2010], where most moisture is concentrated. This provides a simple framework for predicting changes in $P - E$ with warming: wet regions will get wetter and dry regions will get drier, with a fractional change determined by Clausius-Clapeyron. It presupposes that the existing circulations simply transport more moisture. This framework works best for large spatial averages, for which circulation shifts are relatively unimportant [*Held and Soden*, 2006]. Circulation shifts and the weakening of tropical overturning circulations can offset part of the thermodynamic change in $P - E$ [*Chou and Neelin*, 2004; *Vecchi and Soden*, 2007; *Chou et al.*, 2009; *Xie et al.*, 2010].

Most of the success of the *wet gets wetter* mechanism comes from its applicability to zonal-mean $P - E$, because zonal-mean circulation features such as the Hadley circulation and storm tracks have a limited response to climate change [*Held and Hou*, 1980; *Schneider*, 2006; *Walker and Schneider*, 2006; *Schneider et al.*, 2010]. However, the remaining zonally anomalous $P - E$, obtained by subtracting the zonal mean, accounts for 60% of the total spatial variance of $P - E$ in the modern climate [*Wills and Schneider*, 2015]. Zonally anomalous $P - E$ is governed largely by divergent stationary-eddy circulations acting on the boundary layer specific humidity [*Wills and Schneider*, 2015, 2016]. Regional dynamic $P - E$ changes can thus result from changes in the location and strength of stationary-eddy convergence/divergence zones with climate change [*Chou and Neelin*, 2004; *Seager et al.*, 2010; *Xie et al.*, 2010; *Chadwick et al.*, 2013]. Here we analyze the extent to which the *wet gets wetter* mechanism applies to changes in zonally anomalous $P - E$. We analyze the zonal variance of $P - E$, which provides a bulk measure of the amplitude of zonal hydrological cycle variations. The zonal variance of $P - E$ is unaffected by zonal shifts of stationary-eddy circulations and should scale with atmospheric moisture better than $P - E$ at a grid point. To the extent it does not, it implies a weakening of stationary-eddy circulations and/or a reduction of zonal $P - E$ variance by transient-eddy moisture flux changes.

2. Changes in the Zonally Anomalous Hydrological Cycle

We analyze the change of $P-E$ in the Coupled Model Intercomparison Project phase 5 (CMIP5), from the *PAST* (1976–2005 in the historical simulations) to the *FUTURE* (2070–2099 in the Representative Concentration Pathways RCP8.5 emission scenario), pooling all 23 models for which the flow fields necessary for our analysis are available (Table S1 in the supporting information). The annual-mean climatology of $P-E$ in the *PAST* exhibits familiar features such as the general wetness of the tropics and extratropics and dryness of the subtropics (Figure 1a). The zonally anomalous component, P^*-E^*, (with $()^* = () - [\,]$, and $[\,]$ the zonal mean) focuses attention on zonally anomalous wet regions such as the Asian monsoon regions, the maritime continent, the South Pacific Convergence Zone, and the Northern Hemisphere storm tracks, and dry regions such as the subtropical lows, the Mediterranean, and the Northern Hemisphere boreal forests (Figure 1b).

There are many $P-E$ changes in the *FUTURE* that are robust across the models (Figure 1c). Some changes, such as the moistening tendency poleward of 45° latitude and the drying tendency in the subtropics, are well produced by a simple thermodynamic scaling based on the fractional change in surface-air specific humidity q_s (Figures 1e and S1) [cf. *Held and Soden*, 2006]:

$$\delta(P-E) \sim \frac{\delta q_s}{q_s}(P-E). \tag{1}$$

Here $\delta()$ is the annual-mean difference of a quantity from the end of the twentieth century (1976–2005) to the end of the 21st century (2070–2099), and all other quantities are annual-mean climatological values in the twentieth century. We use surface-air specific humidity instead of surface-air temperature for simplicity; its changes are largely determined by changes in surface-air temperature because changes in surface-air relative humidity are small over oceans [*Held and Soden*, 2000, 2006; *Schneider et al.*, 2010]. By using annual-mean values in the scaling (1), we ignore seasonal correlations between moisture changes and $P-E$, which are large in the high latitudes of the Northern Hemisphere (beyond 50°N, see Figure S1). This scaling is particularly good for the zonal-mean change, which was the focus of *Held and Soden* [2006]. The regions where it predicts the wrong sign of change (stippling in Figure 1e) arise because subtropical dry zones expand [*Hu and Fu*, 2007; *Lu et al.*, 2007; *Scheff and Frierson*, 2012], because land areas dry [*Byrne and O'Gorman*, 2015], or because tropical circulations shift or weaken [*Chou and Neelin*, 2004; *Vecchi and Soden*, 2007; *Xie et al.*, 2010].

To illustrate the extent to which *wet gets wetter* does not apply to zonally anomalous $P-E$ changes, we show the P^*-E^* change (Figure 1d) and an adaptation of the simple thermodynamic scaling (Figure 1f):

$$\delta\left(P^*-E^*\right) \sim \frac{\delta[q_s]}{[q_s]}\left(P^*-E^*\right). \tag{2}$$

This simple thermodynamic P^*-E^* change is calculated from the fractional change in zonal- and annual-mean surface-air specific humidity $[q_s]$, based on the observation that zonal variations in q_s do not substantially alter the zonally anomalous moisture budget [*Wills and Schneider*, 2015]. Once again, the regions where the sign is incorrect are indicated with stippling, which covers 33% of the globe. Changes in phase and amplitude of stationary-eddy circulations are large enough that many zonally anomalous wet regions get drier, and dry regions get wetter. Most notably, the simple scaling estimate disagrees with modeled changes over most of the tropical oceans, where P^*-E^* changes are largest. However, the zonally anomalous simple scaling gets the sign correct over most land areas. This is because $P-E$ changes over land are generally smaller than in the zonal mean, which reflects the generally weaker $P-E$ climatology over land. These conclusions are unchanged if one modifies the scaling to account for changes in the zonally anomalous surface specific humidity, δq_s^*, by simply taking the stationary-eddy component of (1).

Tropical circulations are well known to change in strength with global warming [*Held and Soden*, 2006; *Vecchi et al.*, 2006]; our analysis shows that the spatial structure is also substantially altered. This is apparent in the weak correlation between the simple themodynamic scaling (equation (2)) and the P^*-E^* change (dashed blue line in Figure 1g), particularly in the tropics. To assess which other mechanisms govern changes in P^*-E^*, we analyze changes in the zonally anomalous moisture budget

$$\delta(P^*-E^*) = \underbrace{-\nabla\cdot\langle \mathbf{u}\delta q\rangle^*}_{\text{(thermo.)}} \underbrace{-\,\nabla\cdot\langle \delta\mathbf{u}\,(q+\delta q)\rangle^*}_{\text{(dynamic)}} \underbrace{-\,\delta\left(\nabla\cdot\langle \mathbf{u}'q'\rangle\right)^*}_{\text{(transient)}}, \tag{3}$$

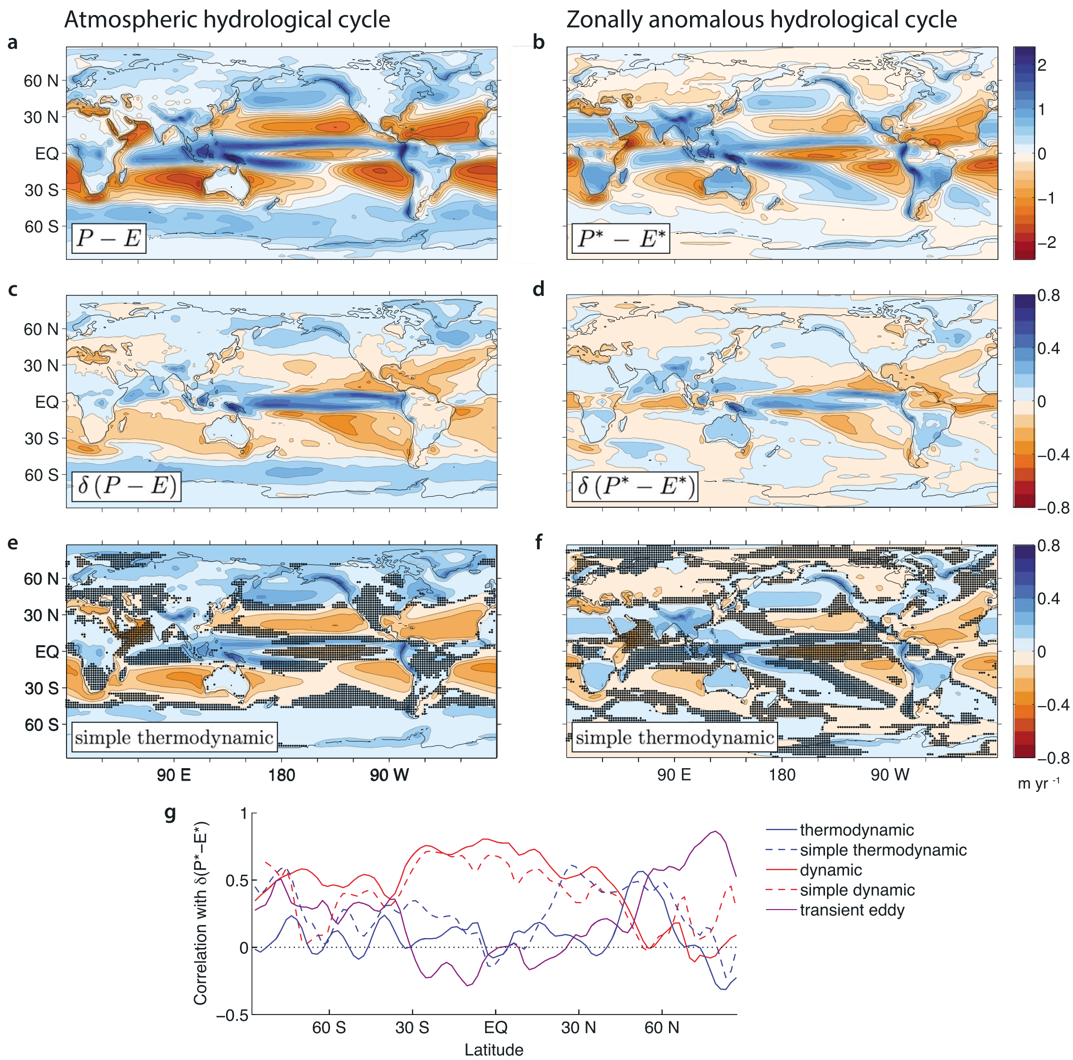

Figure 1. Multimodel mean 1976–2005 climatology of (a) $P - E$ and (b) $P^* - E^*$ and change of (c) $P - E$ and (d) $P^* - E^*$ by the end of the century (2070–2099) in the RCP8.5 scenario. (e and f) Thermodynamic contributions to Figures 1c and 1d, as estimated from the fractional change in surface specific humidity (equations (1) and (2)). Stippling indicates where the thermodynamic scaling is opposite in sign to the simulated change. (g) Zonal correlations of moisture budget changes with $\delta(P^* - E^*)$: thermodynamic, dynamic, and transient-eddy components based on the moisture budget decomposition in equation (3) (solid lines) as well as approximations to the thermodynamic and dynamic components based on equations (2) and (9), respectively (dashed lines).

where $\mathbf{u}$ is the horizontal wind, q is the specific humidity, $()'$ indicates the difference from the annual-mean climatology, and $\langle\rangle$ denotes a mass-weighted vertical integral over the whole domain. The dynamic term here combines the stationary-eddy components of the dynamic term and the nonlinear term of *Seager et al.* [2010], such that the decomposition is exact. The transient-eddy term is computed from a residual and includes all temporal correlations between specific humidity and the horizontal wind, including seasonal correlations [cf. *Wills and Schneider*, 2015]. (The conclusions that follow are not substantially altered by the inclusion of seasonal correlations with the transient-eddy term; see the supporting information.)

Throughout the tropics, subtropics, and Southern Hemisphere midlatitudes, regional $P^* - E^*$ changes are primarily governed by changes in stationary-eddy circulations (red line in Figure 1g). Changes in zonally anomalous transient-eddy moisture fluxes are negatively correlated with $P^* - E^*$ changes in the tropics but play a large role in the extratropical $P^* - E^*$ change in both hemispheres. The thermodynamic term, on which the simple thermodynamic scaling of equation (2) is based, is not well correlated with $\delta(P^* - E^*)$ (Figure 1g). Thus, to understand $P^* - E^*$ changes, one primarily has to understand stationary-eddy changes.

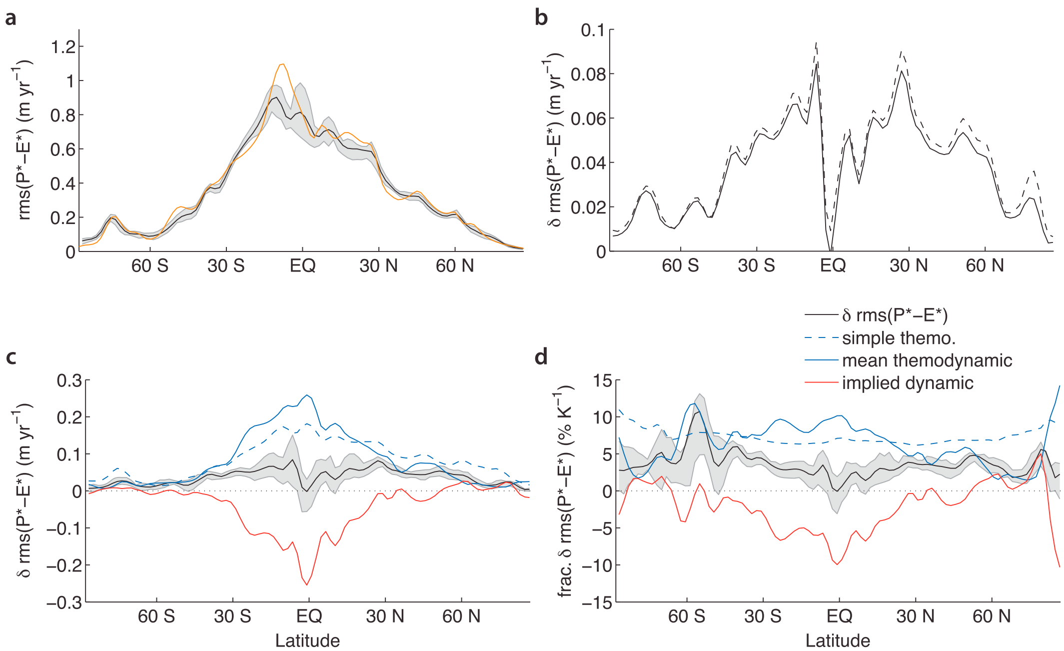

Figure 2. (a) Multimodel mean *PAST* climatology of $\mathrm{rms}(P^* - E^*)$ (black line). Shading shows the interquartile range of the model spread in $\mathrm{rms}(P^* - E^*)$, computed at each latitude. The $\mathrm{rms}(P^* - E^*)$ from ERA-Interim reanalysis (1979–2012) is shown for comparison (orange line) [*Dee et al.*, 2011]. (b) *PAST* − *FUTURE* $\delta\mathrm{rms}(P^* - E^*)$ (solid line) and its approximation by equation (5) (dashed line). (c) Simple thermodynamic (equation (6)) and mean-thermodynamic ($\delta\mathrm{RMS}_{\mathrm{mthermo}}$) estimates of the change in $\mathrm{rms}(P^* - E^*)$. Shading shows the interquartile range of the model spread in $\delta\mathrm{rms}(P^* - E^*)$. Also shown is an implied dynamic contribution based on the difference between the mean-thermodynamic term and the actual change. (d) Fractional change in $\mathrm{rms}(P^* - E^*)$ per degree warming of the zonal-mean surface-air temperature and the contributions from the moisture budget variance terms shown in Figure 2c. Shading shows the interquartile range of the model spread in fractional $\mathrm{rms}(P^* - E^*)$ change per degree warming.

3. Zonal Variance of *P* − *E*

The increased moisture content of the atmosphere can still lead to an increase in the amplitude of zonal variations of the hydrological cycle, characterized by the root zonal variance of $P - E$,

$$\mathrm{rms}(P^* - E^*) \equiv \left[\overline{(P^* - E^*)^2}\right]^{1/2}. \tag{4}$$

The climatology of $\mathrm{rms}(P^* - E^*)$ in the *PAST* simulations, averaged over the 23 CMIP5 models, is generally consistent with $\mathrm{rms}(P^* - E^*)$ from ERA-Interim reanalysis (Figure 2a). The zonally anomalous hydrological cycle, as measured by $\mathrm{rms}(P^* - E^*)$, is strongest in the tropics, where the moisture content of the atmosphere is greatest and the atmospheric circulation is most divergent. It is stronger in the Northern Hemisphere midlatitudes (beyond 45° latitude) than in the Southern Hemisphere midlatitudes because Earth's surface is more zonally inhomogeneous in the Northern Hemisphere. There is an increase in $\mathrm{rms}(P^* - E^*)$ at all latitudes (except directly on the equator) with global warming (Figure 2c). The ensemble-mean increase is greater than the intermodel spread, characterized by the interquartile range of the $\mathrm{rms}(P^* - E^*)$ change (grey shading in Figures 2c and 2d). The fractional change per degree warming of the zonal and annual-mean surface temperature ranges from 0 to 6% K^{-1} (Figure 2d), except at 55°S where the climatological $\mathrm{rms}(P^* - E^*)$ and zonal-mean temperature change are both small.

The change in $\mathrm{rms}(P^* - E^*)$ can be split into thermodynamic, dynamic, and transient-eddy components based on the $P^* - E^*$ variance budget, using the approximate relation,

$$\delta\mathrm{rms}(P^* - E^*) \approx \frac{\delta\left[\overline{(P^* - E^*)^2}\right]}{2 \cdot \mathrm{rms}(P^* - E^*)}, \tag{5}$$

derived in section 6 (Methods). The quality of this approximation is excellent (Figure 2b).

A thermodynamic scaling for rms($P^* - E^*$) is obtained by substituting the thermodynamic scaling for $P^* - E^*$ (equation (2)) into equation (5) and neglecting a further term, which is nonlinear in $\delta[q_s]/[q_s]$, yielding

$$\delta\text{rms}(P^* - E^*) \sim \frac{\delta[q_s]}{[q_s]}\text{rms}(P^* - E^*). \tag{6}$$

This is shown as a dashed blue line in Figures 2c and 2d. The fractional change is given by $\delta[q_s]/[q_s]$ and is approximately 7% K^{-1} at all latitudes equatorward of 70°. The change in rms($P^* - E^*$) is robustly less than this thermodynamic scaling suggests, except for the Southern Ocean, where the variance is small to begin with. We will explore the dynamic changes making up the difference between this thermodynamic scaling and the modeled rms($P^* - E^*$) change. But first, it is beneficial to derive a more accurate estimate of thermodynamic changes.

Following the decomposition of the moisture budget (equation (3)) and the approximation of δrms($P^* - E^*$) (equation (5)), we can split δrms($P^* - E^*$) into thermodynamic, dynamic, and transient-eddy contributions,

$$\delta\text{rms}(P^* - E^*) \approx \delta\text{RMS}_{\text{mthermo}} + \delta\text{RMS}_{\text{mdyn}} + \delta\text{RMS}_{\text{trans}}. \tag{7}$$

The three terms on the right-hand side, which we refer to as the mean-thermodynamic, mean-dynamic, and transient-eddy terms, are given in section 6 (Methods). They include terms that are nonlinear in changes, such that equation (7) is only approximate due to the approximation in equation (5), relating change in rms to changes in variance.

The mean-thermodynamic term, $\delta\text{RMS}_{\text{mthermo}}$, is shown as a solid blue line in Figures 2c and 2d. It differs from the simpler thermodynamic term (6) primarily because it is derived from the time-mean flow contribution to $P^* - E^*$, as opposed to the full $P^* - E^*$, which includes effects of transient eddies (see Figure S2). We will treat transient eddies separately, eventually coming to the conclusion that transient-eddy moisture fluxes do not change thermodynamically. The mean-thermodynamic change is greater than the actual change except in some places at high latitudes. This implies a dynamic change, which is shown as a red line in Figures 2c and 2d. The implied dynamic change is concentrated in the tropics and subtropics. The next section explores the stationary- and transient-eddy components of this dynamic change.

4. Strength of Stationary-Eddy Circulations

In explaining dynamic factors limiting the increase in variance of the zonally anomalous hydrological cycle, we focus, in particular, on the stationary-eddy vertical (pressure) velocities at 850 hPa, ω^*_{850}. This is based on the findings of *Wills and Schneider* [2015, 2016], who show that $P^* - E^*$ can be approximated by

$$P^* - E^* \approx -g^{-1}[q_s]\omega^*_{850}. \tag{8}$$

The intuition behind this approximation is that divergent stationary-eddy circulations carry moisture from the boundary layer to a mean condensation height at about 850 hPa. Additionally, there is a large cancelation between the horizontal advection of moisture by stationary eddies and the transient-eddy moisture flux divergence, as transient eddies relax horizontal moisture gradients set up by horizontal advection. A large portion of the change in $P^* - E^*$ is thus explained by changes in ω^*_{850},

$$\delta(P^* - E^*) \approx -g^{-1}\left([q_s] + \delta[q_s]\right)\delta\omega^*_{850}. \tag{9}$$

The correlation between this simple dynamic change and the full $P^* - E^*$ change is greater than 50% throughout most of the tropics and subtropics (dashed red line in Figure 1g), indicating a close relationship between changes in stationary-eddy vertical motion and changes in $P^* - E^*$ [cf. *Wills and Schneider*, 2016].

From the approximation (8), we can also infer changes in rms($P^* - E^*$),

$$\delta\text{rms}(P^* - E^*) \approx g^{-1}\delta\left([q_s]\text{rms}\left(\omega^*_{850}\right)\right), \tag{10}$$

as was shown in idealized general circulation model experiments [*Wills and Schneider*, 2016]. Here the root zonal variance of stationary-eddy vertical motion, rms(ω^*_{850}), measures the strength of stationary-eddy overturning. The climatology of rms(ω^*_{850}) in the *PAST* is highest in the tropics and subtropics (Figure 3a),

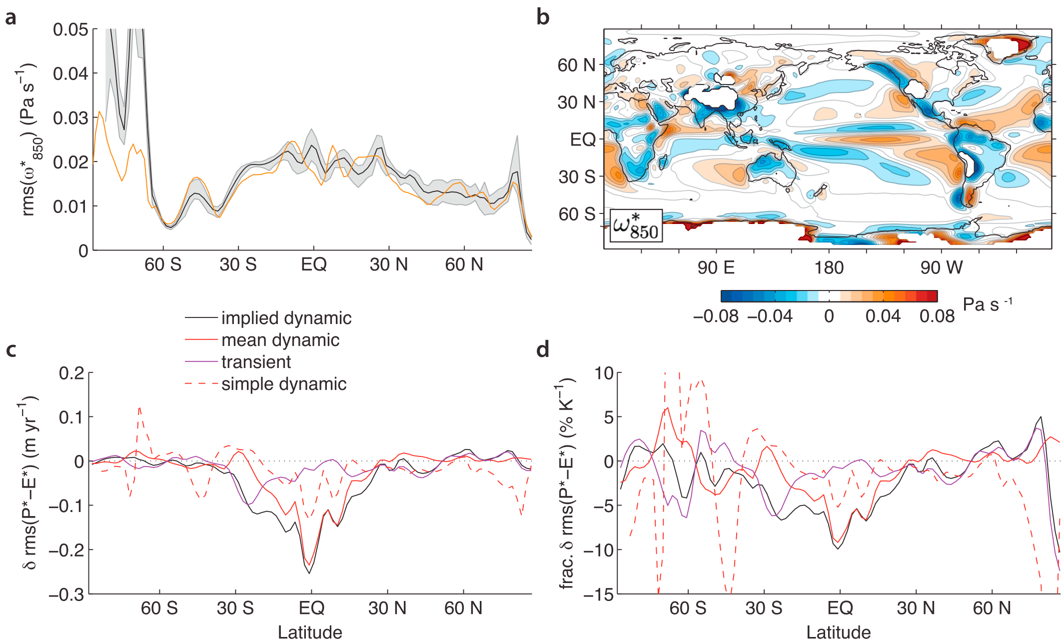

Figure 3. (a) Multimodel mean *PAST* climatology of rms(ω^*_{850}) (black line). Shading shows the interquartile range of the model spread in rms(ω^*_{850}), computed at each latitude. The rms(ω^*_{850}) from ERA-Interim reanalysis (1979–2012) is shown for comparison (orange line) [*Dee et al.*, 2011]. (b) Multimodel mean *PAST* climatology of ω^*_{850}, smoothed with a 200 km Gaussian filter to reduce grid-scale noise. Note the similarity to the $P^* - E^*$ climatology (Figure 1b). (c) Stationary- and transient-eddy contributions to changes in rms($P^* - E^*$), $\delta \mathrm{RMS}_{\mathrm{mdyn}}$, and $\delta \mathrm{RMS}_{\mathrm{trans}}$. These add up to the implied dynamic change (red line in Figure 2c). Also shown is a simple estimate of the dynamic change (equation (11)). (d) Contributions of the moisture budget variance terms shown in Figure 3c to the fractional change in rms($P^* - E^*$) per degree warming of the zonal-mean surface-air temperature.

where divergent circulations are associated with tropical deep convection. It has secondary maxima at high latitudes stemming from topographic vertical motions around Greenland, Patagonia, and Antarctica. Note the similarity between the ω^*_{850} climatology (Figure 3b) and the $P^* - E^*$ climatology (Figure 1b), which emphasizes the relationship between $P^* - E^*$ and ω^*_{850} (8).

Figures 3c and 3d show how dynamic changes contribute to the change in rms($P^* - E^*$). The stationary-eddy dynamic component ($\delta \mathrm{RMS}_{\mathrm{mdyn}}$, solid red lines) is the leading contribution, reducing the zonal variance of $P^* - E^*$, especially in the tropics. The stationary-eddy dynamic change can be approximated from the change in ω^*_{850} variance as

$$\delta \mathrm{rms}(P^* - E^*) \approx g^{-1} \left([q_s] + \delta[q_s]\right) \delta \mathrm{rms}\left(\omega^*_{850}\right), \qquad (11)$$

which includes the product of moisture changes and rms(ω^*_{850}) changes, consistent with the treatment of the nonlinear term in (3). This simple dynamic change shows a slowdown of divergent stationary-eddy circulations (dashed red lines in Figures 3c and 3d). It is a major component of the mean-dynamic change (red lines). This is consistent with the weakening of tropical overturning circulations with global warming as required by energetic constraints on global-mean precipitation [*Betts*, 1998; *Held and Soden*, 2006; *Vecchi and Soden*, 2007; *Schneider et al.*, 2010]. The noisiness of the approximation (11) at high latitudes comes from noisy vertical velocities around Patagonia, Antarctica, and Greenland (see Figure S6). It can be improved by considering zonal anomalies of q_s, which are large over topography. This relation (11) is approximate because it neglects changes in vertical structure of the atmosphere and changes in stationary-eddy moisture advection, and because of the approximation of the climatological $P^* - E^*$ by (8).

Changes in the zonal variance of transient-eddy moisture flux ($\delta \mathrm{RMS}_{\mathrm{trans}}$, purple lines in Figures 3c and 3d) are also important. Transient-eddy moisture flux changes reduce the zonal variation of the hydrological cycle in the tropics and subtropics, but they can increase the zonal variation locally in high latitudes. This corresponds to an amplification of the transient-eddy contribution to the rms($P^* - E^*$) climatology, based on a scaling up of transient-eddy moisture fluxes by moisture changes. This is comparable to their role in changes of the zonal-mean hydrological cycle [*Held and Soden*, 2006; *Wu et al.*, 2011; *Byrne and O'Gorman*, 2015].

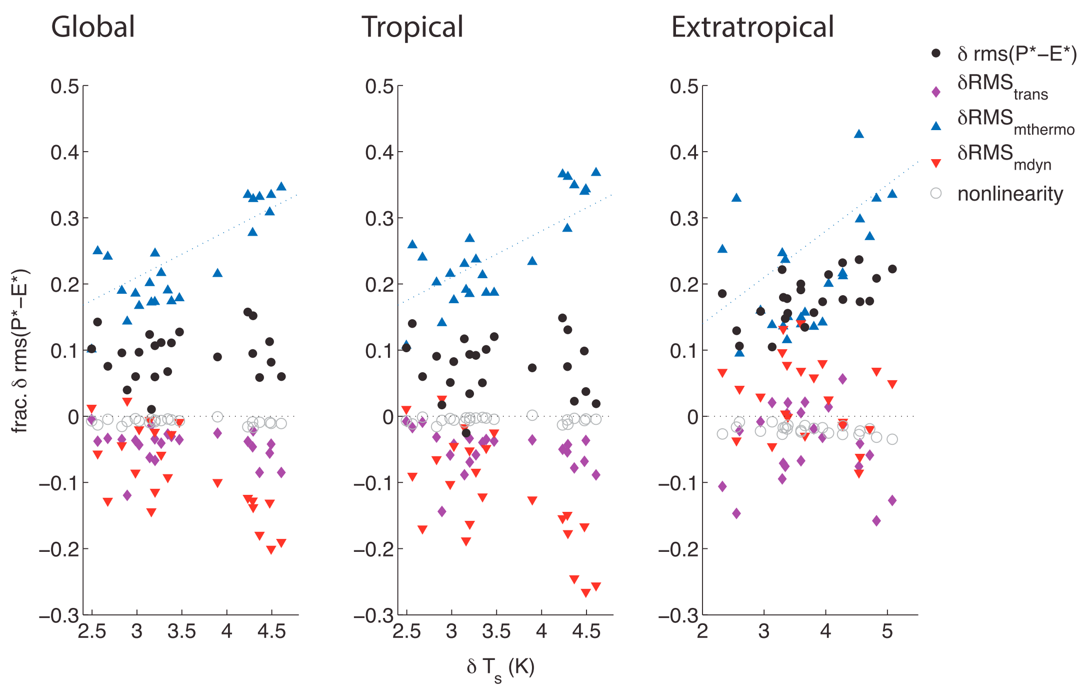

Figure 4. Global-mean, tropical-mean (±0 − 30°), and extratropical-mean (±30 − 75°) fractional change in strength of the zonally anomalous hydrological cycle, $\delta\mathrm{rms}(P^* - E^*)/\mathrm{rms}(P^* - E^*)$ for each model (black dots), plotted versus the *PAST* − *FUTURE* difference in surface-air temperature, averaged over the same region. The mean-thermodynamic, mean-dynamic, and transient contributions to this change are shown separately in blue triangles, red triangles, and purple diamonds, respectively. There is also a contribution from nonlinear (large-amplitude) changes (grey circles). The dotted blue line shows 7% K^{-1} for reference.

The lack of a robust positive contribution from transient eddies in the midlatitudes results from the correlation of transient-eddy moisture flux changes with changes in stationary-eddy moisture flux: transient eddies act diffusively, reducing the $P - E$ variation set up by stationary eddies, leading to a negative contribution to $\mathrm{rms}(P^* - E^*)$. The transient-eddy change is a mixture of changes in moisture content and changes in transient-eddy winds [*Wu et al.*, 2011; *Byrne and O'Gorman*, 2015], but there is no exact means of splitting these contributions. Together, the reduction of stationary-eddy vertical motions and changes in transient-eddy moisture fluxes modify the response of $\mathrm{rms}(P^* - E^*)$ to climate change at all latitudes, but especially in the tropics.

Given the intermodel spread in zonal-mean $P-E$ [see, e.g., *Voigt and Shaw*, 2015], it would be surprising if the models agreed on the change of $\mathrm{rms}(P^* - E^*)$. To investigate the intermodel spread in $\mathrm{rms}(P^* - E^*)$ change, we show the global-mean, tropical-mean, and extratropical-mean fractional $\mathrm{rms}(P^* - E^*)$ change for all models in Figure 4. It is plotted against the change in surface-air temperature, averaged over the same region, with the expectation that the intermodel spread in climate sensitivity would explain some of the intermodel spread in $\delta\mathrm{rms}(P^* - E^*)$. All models show an increase in global-mean $\mathrm{rms}(P^* - E^*)$ with warming. This change is split into mean-thermodynamic, mean-dynamic, and transient-eddy components (blue triangles, red triangles, and purple diamonds) as well as a component (grey circles) due to large-amplitude changes neglected in the approximation (5).

In all models, global-mean $\mathrm{rms}(P^* - E^*)$ increases, but less than would be expected from thermodynamics alone, implying a moderating contribution from dynamic changes. The majority of models show a larger contribution from stationary-eddy than transient-eddy dynamic changes. The fractional change in $\mathrm{rms}(P^* - E^*)$ is somewhat larger in the extratropics than the tropics. In the extratropics, the mean-thermodynamic term is more scattered due to differences in the stationary-eddy moisture flux convergence climatology among models. The extratropical mean-dynamic change can either be positive or negative depending on the model, but the extratropical transient-eddy change is predominantly negative or weakly positive.

Transient-eddy moisture flux changes decrease the zonal variance of the hydrological cycle in the extratropics. This is contrary to the expectation based on the simple thermodynamic scaling and is thus primarily a dynamic change resulting from the diffusive nature of transient-eddy moisture fluxes acting on zonal anomalies set by stationary eddies.

5. Conclusions

The CMIP5 models studied show a robust strengthening of the zonally anomalous hydrological cycle with warming in the RCP8.5 scenario. The strengthening comes about primarily from the increasing moisture content of the atmosphere. However, it is smaller than expected from various estimates of the thermodynamic change implied by the increased atmospheric moisture content, because dynamic changes reduce the zonal variance of the hydrological cycle in the global mean. In the tropics, a weakening of divergent stationary-eddy circulations dominates the dynamic change. Transient-eddy moisture flux changes dominate dynamic changes in the extratropics, decreasing the midlatitude zonal variance response on average. This analysis, applied seasonally leads to similar conclusions: the zonal variance of $P - E$ increases with global warming in all seasons, but thermodynamic changes are largely counteracted by a weakening of stationary-eddy overturning (see supporting information).

These results imply that it is important to understand stationary-eddy vertical velocities in order to understand the zonally anomalous hydrological cycle. In the tropics, there are some constraints on stationary-eddy vertical velocity changes from energetic and moist static energy arguments [*Betts*, 1998; *Chou and Neelin*, 2004; *Held and Soden*, 2006]. Understanding stationary-eddy vertical velocities in the subtropics and midlatitudes remains an open challenge.

Understanding changes in strength of the zonally anomalous hydrological cycle is one aspect of a complete understanding of the response of regional $P - E$ to climate change. The zonal-mean $P - E$ response is approximately described by the *wet gets wetter, dry gets drier* mechanism, with some influence of changes in the Hadley circulation, shifts in storm tracks, and relative humidity and moisture gradient changes [*Held and Soden*, 2006; *Byrne and O'Gorman*, 2015]. Understanding changes in strength of the zonally anomalous hydrological cycle provides the next order understanding. The small strengthening means that areas that are wetter than the zonal mean (Southeast Asia, Oceania, and the west coast of the Americas) may get somewhat wetter and areas that are drier than the zonal mean (continental Asia and subtropical ocean highs) may get somewhat drier, though with fractional changes significantly smaller than the 7% K^{-1} predicted from the Clausius-Clapeyron relation. Shifts in stationary-eddy circulations exert an additional influence on regional $P - E$ changes, and they should be the focus of future work.

6. Methods

6.1. Spatial Variance Budget

We analyze changes in RMS $\equiv$ rms($P^* - E^*$) by relating them to changes in the spatial variance of $P^* - E^*$, which can be expanded as

$$\begin{aligned} \delta\left[(P^* - E^*)^2\right] &= \delta(\mathrm{RMS}^2) \\ &= (\mathrm{RMS} + \delta\mathrm{RMS})^2 - \mathrm{RMS}^2 \\ &= 2 \cdot \delta\mathrm{RMS} \cdot \mathrm{RMS} + (\delta\mathrm{RMS})^2. \end{aligned} \quad (12)$$

The second quadratic term is small and can be neglected. The result is an expression for δRMS in terms of $\delta[(P^* - E^*)^2]$,

$$\delta\mathrm{RMS} \approx \frac{\delta[(P^* - E^*)^2]}{2 \cdot \mathrm{RMS}}, \quad (13)$$

where the quadratic nonlinearity has been removed. The change in $P^* - E^*$ variance is given by

$$\delta\left[(P^* - E^*)^2\right] = 2\left[(P^* - E^*)\delta(P^* - E^*)\right] + \left[\delta(P^* - E^*)^2\right]. \quad (14)$$

The second term, which is nonlinear in changes, is included in the analysis but is generally much smaller than the first term.

We split the change in $\delta\mathrm{rms}(P^* - E^*)$ into thermodynamic, dynamic, and transient-eddy components based on

$$\begin{aligned}\delta(P^* - E^*) &= -\nabla \cdot \langle \mathbf{u}\delta q\rangle^* - \nabla \cdot \langle \delta\mathbf{u}\,(q+\delta q)\rangle^* - \delta\left(\nabla \cdot \langle \mathbf{u}' q'\rangle\right)^* \\ &\equiv \quad \Delta_{\mathrm{thermo}} \quad + \quad \Delta_{\mathrm{dyn}} \quad + \quad \Delta_{\mathrm{trans}}.\end{aligned} \tag{15}$$

The transient-eddy term Δ_{trans} is diagnosed as a residual. Substituting this decomposition into equations (13) and (14), we obtain

$$\delta\mathrm{RMS} \approx \delta\mathrm{RMS}_{\mathrm{mthermo}} + \delta\mathrm{RMS}_{\mathrm{mdyn}} + \delta\mathrm{RMS}_{\mathrm{trans}}, \tag{16}$$

where

$$\begin{aligned}\delta\mathrm{RMS}_{\mathrm{mthermo}} &\equiv \frac{1}{2}\frac{1}{\mathrm{RMS}}\left[2(P^* - E^*)\Delta_{\mathrm{thermo}} + \Delta^2_{\mathrm{thermo}}\right], \\ \delta\mathrm{RMS}_{\mathrm{mdyn}} &\equiv \frac{1}{2}\frac{1}{\mathrm{RMS}}\left[2(P^* - E^*)\Delta_{\mathrm{dyn}} + \Delta^2_{\mathrm{dyn}} + 2\Delta_{\mathrm{dyn}}\Delta_{\mathrm{thermo}}\right], \\ &\text{and} \\ \delta\mathrm{RMS}_{\mathrm{trans}} &\equiv \frac{1}{2}\frac{1}{\mathrm{RMS}}\left[2(P^* - E^*)\Delta_{\mathrm{trans}} + \Delta^2_{\mathrm{trans}} + 2\Delta_{\mathrm{trans}}\left(\Delta_{\mathrm{thermo}} + \Delta_{\mathrm{dyn}}\right)\right].\end{aligned} \tag{17}$$

All fields are smoothed with a 200 km Gaussian filter before any variance statistics are computed. The motivation for this smoothing is the large amount of grid-scale noise present in the unprocessed fields for some models, especially for vertical velocities (see Figure S6).

The quadratic terms in (17) are generally small, but significant. Therefore, the changes in RMS can be understood as being proportional to the zonal correlation of Δ_{trans}, Δ_{trans}, and Δ_{trans} with $P^* - E^*$, but we retain the quadratic terms for completeness. The philosophy behind the arrangement of quadratic terms is that $\delta\mathrm{RMS}_{\mathrm{mthermo}}$ should be the change expected with no knowledge of what happens to the dynamic terms, that $\delta\mathrm{RMS}_{\mathrm{mdyn}}$ should be the improvement by including stationary-eddy changes with no knowledge of transient-eddy changes, and that $\delta\mathrm{RMS}_{\mathrm{trans}}$ should be everything else. Note that $\delta\mathrm{RMS}_{\mathrm{trans}}$ in the extratropics (Figures 3 and 4) is the residual of positive contributions from $[2(P^* - E^*)\Delta_{\mathrm{trans}} + \Delta^2_{\mathrm{trans}}]$ and negative contributions from $[2\Delta_{\mathrm{trans}}(\Delta_{\mathrm{thermo}} + \Delta_{\mathrm{dyn}})]$.

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